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REGIONAL TECTONIC ANALYSIS OF VENUS EQUATORIAL HIGHLANDS AND
COMPARISON WITH EARTH-BASED AND MAGELLAN RADAR IMAGES

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In 1980, Pioneer Venus showed the surface of Venus to be very different from that of the Earth. The new radar images from Magellan have shown Venus to be even more active, complex, and diverse than the original Pioneer Venus and Venera data indicated (Saunders et al., 1991). Most notably, very little has been seen on Venus that has direct analogies to Earth, including a global tectonism which shows no evidence of plate tectonics (Solomon et al., 1991) as was first inferred from the Pioneer Venus altimetry data (e.g. Phillips et al., 1981). The mechanisms by which heat is removed from Venus are not known, but probably include some combination of conduction through areas of thin lithosphere and extrusion of magma onto the surface. A related question involves the mechanism by which topography on the planet is compensated. Some areas show apparent depths of compensation of 100 km or greater (Sjogren et al., 1983; Kaula, 1984; Bills and Kiefer, 1985; Grimm and Phillips, 1992). Other areas exhibit shallow compensation (Sjogren, 1981; Williams and Gaddis, 1991). The surface temperature of 450°C extrapolated downward for any realistic geotherm would indicate high temperatures at shallow depths, making support due to a thick lithosphere or deep crustal roots unlikely. The deep compensation is probably an indicator that dynamic compensation, driven by upwelling and downwelling plumes in the mantle, is largely responsible for topographic support. Ultimately, the source of this compensation will be understood by understanding the characteristics of the mantle of Venus, such as the temperature distribution, bulk chemical makeup, volatile content, and patterns of convection, both past and present.

Over the past two years, the tectonic analysis of Venus we have done has centered on global properties of the planet, in order to understand fundamental aspects of the dynamics of the mantle and lithosphere of Venus. These include studies pertaining to the original constitutive and thermal character of the planet, as well as the evolution of Venus through time, and the present day tectonics. Recent work by a number of groups has indicated that details of the thermal history 500 million to 1 billion years ago may have profound implications to the present day tectonic character of Venus (Phillips et al., 1992; Schaber et al., 1992; Solomon, 1993). The cratering record of Venus indicates the possibility of a single

large resurfacing event or a change in the character of tectonism at this time.

We have developed parameterized convection models of the Earth and Venus. Parameterized convection models treat the mantle as a heat engine, which produces heat through the radioactive decay and loses heat through the surface of the planet. Convection in the mantle regulates the heat balance in the interior, and these processes can be modelled numerically. These models assume whole mantle internally-heated convection characterizes the mantles of both planets, unlike earlier models for Earth which assumed convection heated from below. The viscosity is temperature, volatile-content, and stress dependent. An initial temperature and volatile content is assumed, and the thermal evolution is tracked for 4.6 billion years. During this time, heating occurs by decay of radiogenic elements in the mantle, and degassing and regassing of volatiles takes place at the surface. The initial complement of radiogenic material is set by the requirement that the present day heat flow be 70 mW/m^2 for the Earth. For a model assuming plate tectonics as the primary heat loss mechanism (Fig. 1), representing the Earth through most of its history and perhaps Venus' earlier history, degassing of the mantle was found to occur rapidly ($\sim 200 \text{ My}$) over a large range of parameters (Williams and Pan, 1990b, 1992). Even for parameters chosen to represent extreme cases of an initially cool planet, low radiogenic heating, and large initial volatile complement, the mantle water content was degassed to an equilibrium value in about 2 By. We were also able to show that these results were not affected by assuming a weakly stress dependent viscosity or by the degassing and regassing efficiencies assumed for the Earth. We also calculated the radiogenic Ar^{40} degassed from the mantle through time. Argon 40 is produced through time by the slow radiogenic decay of K^{40} , so the evolution of Ar^{40} at the surface is not as sensitive to initial conditions as the initial volatiles, such as water. We found that the degassing efficiency assumed had a substantial effect on the total outgassed to the present. The general conclusion of this work, however, is that for an initially hot Earth, water was outgassed from the mantle rapidly and reached equilibrium surface concentrations in about 200 My for all models, regardless of the other conditions assumed. This would indicate that the

present mantle water budget is currently in equilibrium with the surface water, or may even be increasing with time as the mantle cools and outgasses less efficiently.

We noted that cooler initial values were also tried in our Earth parameterized convection models, and that these models showed water taking a longer time to be outgassed from the mantle. These values may be applicable to the early Venus, if a large, Moon-forming impact on Earth resulted in efficient heating and loss of water (Kaula, 1990), while such a large impact was absent on Venus. This may be unlikely, according to the models of Wetherill (1991), but the lack of a moon on Venus and the slow retrograde rotation could be indicators that the large impacts on Venus were smaller than Earth's Moon-forming impact. If Venus did avoid such a giant impact, it would leave it with a comparably greater volatile budget, lower initial temperatures, and less vigorous early convection. It would still be impossible to retain large amounts of water in the interior of Venus for 4.6 billion years unless another mechanism could operate at cooler temperatures to trap the water in the interior. We have postulated such a mechanism (Williams and Pan, 1990a). This mechanism traps crust-forming melts within the mantle due to a cusp in the water-saturated solidus of olivine (Fig. 2), causing these melts to refreeze at depth into a dense eclogite phase, which will inhibit ascent of this material to the surface. As a packet of hotter mantle material rises, it intersects the wet solidus (point X) and begins to melt. The initial melt fraction contains most of the volatiles and would form typical basalt if it reached the surface. At some critical fraction of partial melt, the liquid will begin to rise faster than the surrounding solid, and will eventually re-intersect the solidus at point Y. At this point, the melt will resolidify. Because this resolidification takes place below the basalt/eclogite transition (Z) the melt will freeze to become eclogite. Eclogite is denser than the surrounding olivine, so it will tend to sink, preventing the volatiles from reaching the surface. This effect, however, requires a hydrous mantle so early loss of water might prevent it from taking place. The viability of this mechanism therefore depends strongly on details of the early history of the planet. Since without plate tectonics there is no mechanism for regassing volatiles into the mantle of

Venus, as occurs on Earth at subduction zones, this means the interior of Venus would at present be almost completely dry.

We have reformulated our parameterized convection code to model Venus with an initially hydrous mantle to determine how the "cold-trap" could affect the evolution of the planet (Williams and Pan, 1991). Assuming heat and volatiles are lost through extrusion of material onto the surface, this leads to an episodic magmatic history (Fig. 3). During periods of no magmatism when the cold-trap is in effect, the inefficient loss of heat due to conduction alone causes the temperature to rise. At a high enough temperature, the cold-trap no longer operates because the geotherm is high enough to avoid the cusp in the solidus, and magma will reach the surface in large volumes. This leads, at least locally, to an efficient loss of heat and volatiles from the interior. Cooling of the interior causes the cold-trap to take effect, and the cycle begins again. This effect could take place on a regional scale, but would probably not be globally simultaneous. It therefore could not explain the postulated global resurfacing event of Schaber et al. (1992), but could be an explanation of the episodic nature of regional large-scale volcanism seen in the Magellan images (Phillips et al., 1991).

One important parameter needed to understand the effects of stress on surface morphology is the crustal thickness. For our work on Tellus Regio (Williams and Gaddis, 1991) we were not able to constrain the thickness of the crust. We have attempted to put constraints on this value through the use of crustal extrusion models (Bird and Williams, 1990). As shown in Fig. 4, for a given crustal thickness, surface features of different wavelengths will relax at different rates (Bird, 1991). This effect is due to the fact that as crustal material gets hotter with depth, it tends to get weaker. Below the crust, mantle material tends to be stronger than the crust at similar temperatures, so a weak layer of hot crust is trapped between a strong cool crust and a strong hot mantle. This weak crust will be extruded from beneath high topography due to the overlying burden. Intermediate wavelength features are especially sensitive to crustal thickness, and a thick crust should exhibit a noticeable relaxation of features of these wavelengths. Two-dimensional Fourier analysis was used

to determine the power spectra of the Venus topography for different areas in the equatorial highlands. These spectra showed that no significant wavelength-dependent relaxation has taken place in these areas. Given the high Venus surface temperature, this limits the crustal thickness for reasonable values of heat flow to less than 20 km in the highland regions studied, unless these regions are very young and have not had time to relax. A thin crust in the highland regions of Venus has major implications to compensation models of these regions, as the crust itself cannot be a major contributor to the support of topography unless it is thick enough to form substantial crustal roots.

It is important to know how Venus started out, particularly in terms of volatiles, because the present-day volatile content of the Venus mantle may have important implications to the current behavior of both the mantle and the surface. Volatile content determines in large part both the strength of the surface and lithosphere, and also the type of magmatism expected at the surface of Venus. The rheology of the mantle is particularly sensitive to volatile content. High water content will tend to make the mantle weaker at a given temperature, and hence flow and melt more readily. While we know the surface of Venus is extremely dry, we have very little information on the volatile content on the Venus mantle. The Magellan images provided fairly conclusive evidence that plate tectonics do not operate on Venus and that volcanism was important in the past (Solomon et al., 1991). Distributed magmatism may therefore be an important mode of heat release from the interior, and learning the form taken by this magmatism is critical to our understanding of Venus (see e.g. Phillips et al., 1991).

We have developed models of early planetary embryo collisional and thermal history. Using models developed by Wetherill (1991, 1993) the growth and collisional history of early moon-size planetary embryos has been modelled. These bodies are formed from the initial 5 to 10 km planetesimals in the solar nebula by relatively low velocity impacts on the order of the escape velocity. A number of large runaways form rapidly, and these become the planetary embryos. Using finite-difference thermal models and collisional shock heating calculations, the thermal history of

these bodies is being examined (Williams et al., 1992). These models show that high temperatures can be reached in the interiors of these bodies, even before they start impacting and coalescing into final planetary bodies (Fig 5). This may be important to the early history of volatiles in the solar system, and may have a large influence on the early volatile budgets of the terrestrial planets. Work on this phase and its implications to Venus and Earth is still ongoing.

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FIGURE CAPTIONS

Figure 1: Parameterized convection models of the evolution of mantle water abundance with time. All models have an initial complement of 4 ocean masses (0.00137 wt. fraction water). The number associated with each curve gives the assumed initial homologous temperature (ratio of temperature to melting temperature, T_m). Note that even for a planet starting at only 0.6 T_m , the mantle has degassed its water to an equilibrium value in about 2 billion years (Williams and Pan, 1991b).

Figure 2: Pressure-temperature diagram illustrating the "cold-trap" effect. As material rises along geotherm 1 in a hydrous mantle, it crosses the H₂O-undersaturated solidus at point X and begins to melt. At point Y, the melt recrosses the solidus, due to the hornblende/phlogopite transition, and refreezes. Because this resolidification takes place below the basalt/eclogite transition (Z), the melt solidifies in the form of eclogite, which is denser than the surrounding mantle. This inhibits the rise of melt products, including volatiles, to the surface and thereby inhibits crustal formation. An anhydrous solidus is also shown. Geotherm 2 represents a high temperature case, in which the hydrous cusp is never recrossed, so that melt can reach the surface (Williams and Pan, 1990a).

Figure 3: Thermal and volatile evolution of Venus with a hydrous mantle and "cold-trap" over the last billion years. This diagram assumes Venus has held enough water in its interior through its early history to maintain a hydrous mantle. The homologous temperature (top sawtooth curve) indicates the episodic nature of the thermal evolution of Venus under these circumstances. The mantle heats up until the melting temperature is reached. Then the geotherm no longer intersects the cusp (Fig. 2) and the melt can escape, cooling the mantle and allowing loss of interior water, as indicated in the lower curve (see Williams and Pan, 1991a).

Figure 4: Illustration of how thicker crust will show more significant relaxation with crustal extrusion. This is a diabase crust, varying in thickness for 20-56 km assuming Airy compensation and a temperature gradient of 5 C/km. The top curve shows a cross-section of initial topography. After 1 billion years, the topography has the form shown in the bottom curve. Note that the thicker crust to the right shows significant relaxation of the 100-500 km topography, while the thinner crust to the left exhibits little change (see Bird and Williams, 1990).

Figure 5: Thermal evolution of an accreting planetary embryo. The maximum temperature inside the body is shown as a function of time. Large impacts are obvious by the sudden jump in temperature. The accretion and thermal history are calculated as in Williams et al., 1992. Note that the final temperatures reached by the embryo are strong functions of the last large impacts.

Figure 1

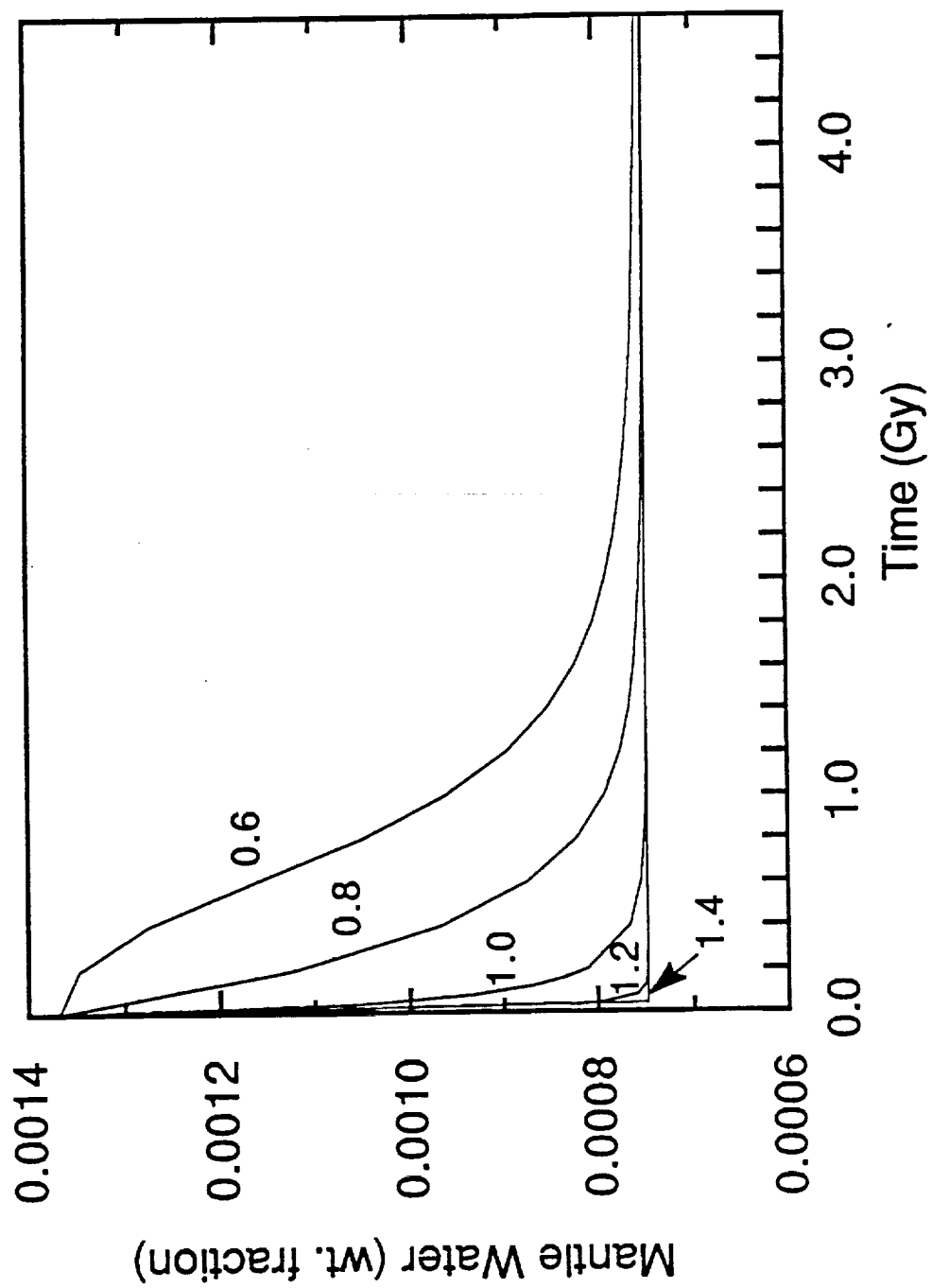


Figure 2

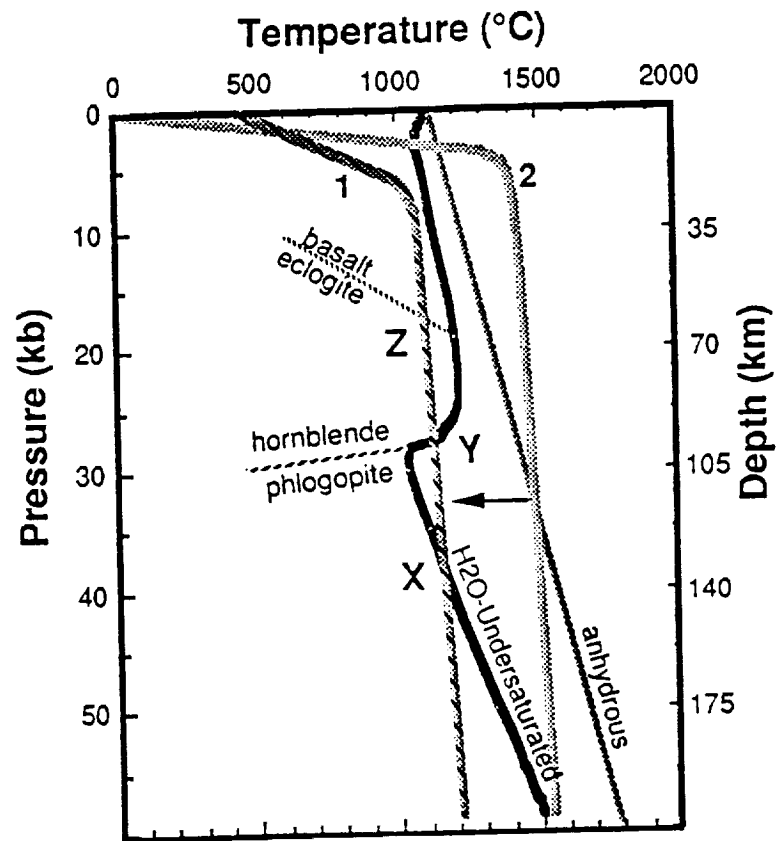


Figure 3

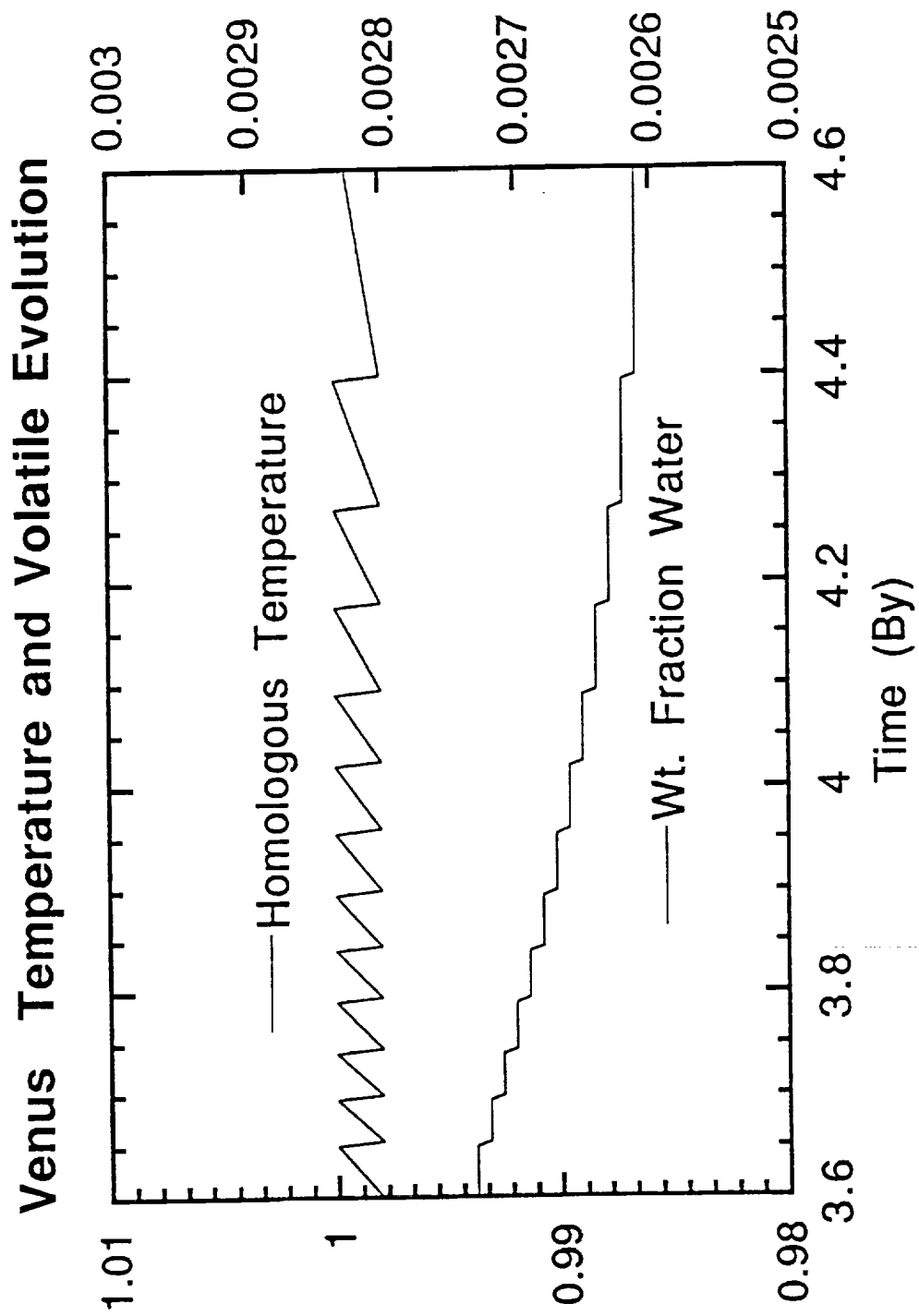


Figure 4

AIRY COMPENSATION, 20-56 KM CRUST OF DIABASE, 5 C/KM

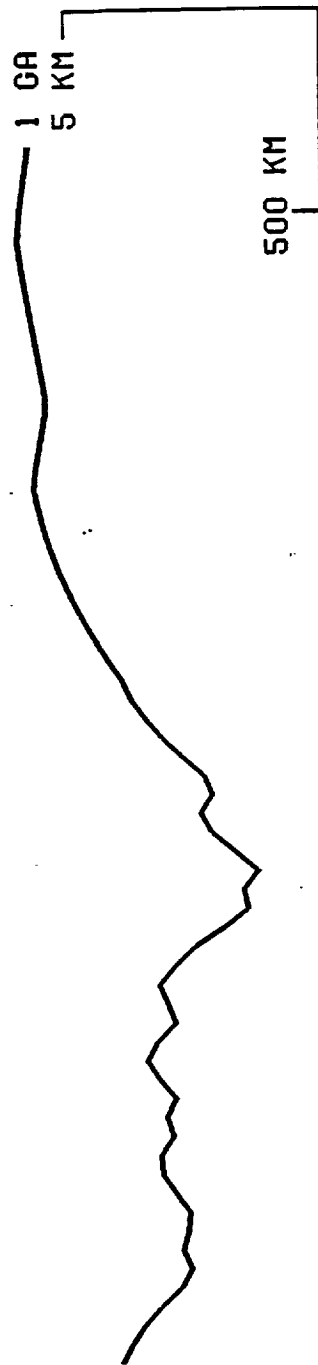
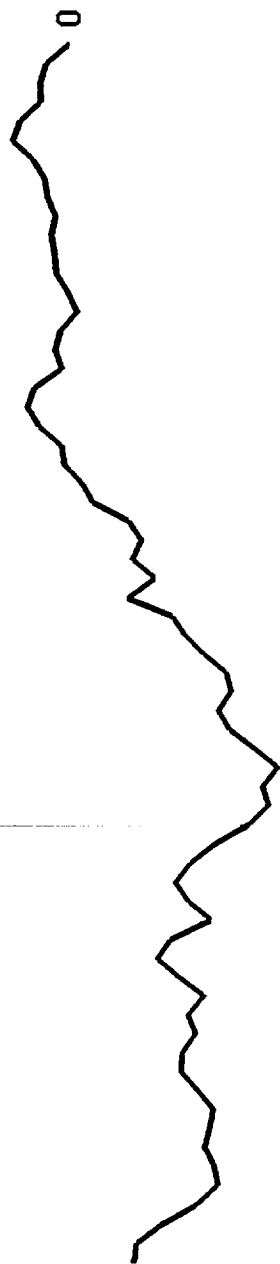


Figure 5

Planetary Embryo Evolution

